Renewable and Sustainable Energy Reviews 16 (2012) 4264–4276

Contents lists available at SciVerse ScienceDirect

Renewable and Sustainable Energy Reviews

journal homepage: www.elsevier.com/locate/rser

Solar attenuation by aerosols: An overview

Yousef A. Eltbaakh[a,*], M.H. Ruslan[b,*], M.A. Alghoul[b,*], M.Y. Othman[b], K. Sopian[b], T.M. Razykov[b]

[a] Physics Department, Universiti Kebangsaan Malaysia, 43600 Bangi, Selangor, Malaysia
[b] Solar Energy Research Institute, Universiti Kebangsaan Malaysia, 43600 Bangi, Selangor, Malaysia

ARTICLE INFO

Article history:
Received 19 October 2011
Accepted 18 March 2012
Available online 2 May 2012

Keywords:
Solar attenuation
Atmospheric aerosol
Turbidity
Cloudless sky conditions
Ångström coefficients
Linke coefficient

ABSTRACT

One of the most important factors affecting the attenuation of solar radiation reaching the earth's surface under cloudless sky conditions is the presence of aerosol particles. A direct measurement of the aerosol transmittance is not possible due to the strong influence of the other atmospheric components. Thus the extinction caused by the atmospheric aerosol can be calculated only indirectly using fundamental physical techniques. This work provides an overview of the effect of aerosols on solar radiation budget by considering two common turbidity parameters including the Linke turbidity factor T_L and Ångström turbidity coefficients β and α. Total extinction of solar radiation due to the absorption and the scattering caused by the atmospheric aerosol accounts for 10–20% for zero zenith angle. The influence of aerosol on radiation passing through the atmosphere cannot be neglected, especially in urban or industrialized areas. The attenuation of solar radiation through a real atmosphere versus that through a clean dry atmosphere gives an indication of the atmospheric turbidity.

Contents

1. Introduction

One of the most important factors affecting the amount of solar radiation reaching the earth's surface under cloudless sky conditions is the presence of aerosol particles (small particles either in solid state or in liquid state) in the atmosphere [1,2]. The aerosol number or mass concentration is key parameters in the studies of atmospheric radiative effects [3]. Aerosols are present in both troposphere and stratosphere and mostly throughout the atmospheric boundary layer at number concentrations depending upon factors such as location, atmospheric conditions, annual and diurnal cycles and presence of local sources [4]. Natural aerosol particles range in radius from 1 to 10^5 nm; very small particles (called Aitken particles) from 1 to 10 nm; and large particles from 10^2 to 10^3 nm. Particles in the 10^3–10^5 nm range are called giant particles [5,6]. According to the formation of the aerosol dispersions, they fall into two categories, viz., dispersed aerosols and condensed aerosol [2]. The presence of aerosols make the atmosphere turbid and the determination of turbidity, the property of the atmosphere in the presence of aerosols is of importance in climatology, in pollution studies and for solar energy utilization [7]. Aerosols influence the solar radiation both directly and indirectly through their various sizes and thus their different optical and physical properties. When aerosols are sufficiently large in size, they scatter and absorb sun light, and when these particles are small, they act as cloud condensation nuclei and aid in the formation of clouds. The attenuation caused by absorption and scattering by aerosols, is known as atmospheric turbidity. The assessments of radiative effects of aerosols, especially of the indirect effect on the radiative balance of the earth are highly uncertain. Uncertainty arises in part, because of the differences in the assumed aerosol concentration and evaluation of the cloud droplet number concentration (CDNC) [3]. Although the amount of aerosols presented in the atmosphere in the vertical direction can be quantified in terms of the number of particles per unit volume [8] or their mass in microgram per cubic meter [7] existing techniques for determining this number is practically applicable only for aerosols presented near the earth's surface [8].

* Corresponding authors.
E-mail addresses: Y.A.Eltbaakh@gmail.com (Y.A. Eltbaakh), hafidz@pkrisc.cc.ukm.my (M.H. Ruslan), dr.alghoul@gmail.com (M.A. Alghoul).

1364-0321/$ – see front matter © 2012 Elsevier Ltd. All rights reserved.
doi:10.1016/j.rser.2012.03.053

Nomenclature

β	the Ångström's turbidity coefficient
α	the Ångström wavelength exponent
N	number density of aerosols
λ	wavelength
τ_λ	aerosol optical depth at the wavelength λ
r	radius of aerosols
ν	fitting parameter characterizing the size distribution
h	solar elevation angle
β_0	true value of Ångström's turbidity coefficient
α_0	true value of Ångström wavelength exponent
$OG1$	old number of cutoff filter OG530, opaque up to 530 nm; transparent 530–2800 nm
$RG2$	old number of cutoff filter RG630, opaque up to 630 nm; transparent 630–2800 nm.
$RG8$	old number of cutoff filter RG695, opaque up to 695 nm; transparent 695–2800 nm
w	water vapor content
T_L	the Linke turbidity factor for an air mass equal to 2
T_{Lm}	long-term monthly average of the Linke turbidity factor for an air mass equal to 2
$\delta_R(m)$	the integral Rayleigh optical thickness
δ_{cda}	clear and dry atmosphere optical thickness
m	air mass
T_{LIP}	Linke turbidity factor corrected by Ineichen and Perez

However, it is more common to represent the amount of aerosols by an index of turbidity [7].

Due to the relationship existing between aerosols and attenuation of solar radiation reaching the earth's surface [9] several so-called "turbidity indices" have been defined during the past decades, differing slightly in physical meaning. The basic relationship underlying the derivation of the various turbidity indices is the Lambert–Bouguer–Beer law [10]. There are at least 11 methods which discussed atmospheric turbidity in the literature such as Linke and Boda [11], Kastrov [12], Ångström [13], Feussner and Dubois [14], Schneider [15], Schuepp [16], Volz [17], Valko [18], Unsworth–Monteith [19], Davies and Hay [20], illuminance turbidity [21]. Among these atmospheric turbidity coefficients, the most frequently used are the Ångström turbidity coefficients and the Linke turbidity factor, as reported by Wen and Yeh [22]. The Ångström exponent is very popular because of the simplicity of the respective equation, because it enables to interpolate or to extrapolate aerosol optical properties, and because it is connected to particle microphysics [23]. In addition, it also indicates the amounts of aerosols and represents the combined effects of both scattering and absorption caused by aerosols. It also indicates the amounts of aerosols [24]. The Linke turbidity factor T_L is of particular importance for the calculation of beam solar and diffuse radiation in a determined locality. This parameter is also useful for the prediction of the availability of solar radiation and day-light illuminance under cloudless skies [25] and it is a key input to several models assessing the downwelling irradiance under clear skies that are used by several communities in the fields of renewable energies, climatology, agro-meteorology, and atmospheric pollution [26].

2. Ångström's turbidity coefficients β and α

The aerosol optical thickness depends not only on aerosol characteristics (size distribution, refractive index, etc.) but also on aerosol total loading. Ångström suggested a single formula for aerosol scattering optical thickness evaluation generally known as Ångström's turbidity formula given by the following [27]

$$\tau_\lambda = \beta\lambda^{-\alpha}$$

In this equation β is called the turbidity coefficient and α the wavelength exponent.

The Ångström turbidity coefficient β represents the amount of aerosols in the atmosphere in the vertical direction [28]. The parameter β, which may vary from 0.0 to 0.5, is an index of the amount of aerosols present in a vertical column of the atmosphere. The parameter α is a reliable index of the size distribution of these aerosols (at least for particles of radii ranging from 0.1 to 1.0 mm), as indicated by the nephelometer measurements of Charlson (1972) cited in [1]. In the solar spectrum, the Ångström exponent α is a good indicator of the dominant size of the atmospheric particles [29]. Fluctuations of α reflect aerosol size distribution variations. The maximum value of α (equal to 4) corresponds to molecular extinction. Values near zero (or even negative) correspond to coarse-mode aerosols (sea spray and desert dust) indicating a non-aerosol optical depth (τ) wavelength dependence, while values of α above 1.5 indicate significant presence of fine-mode particles (mainly smoke or urban aerosols). The average α over the whole spectrum between 290 and 4000 nm has a value between 0.9 and 2.0 with the most frequent observations around $\alpha = 1.3$ [30].

Several methods may be used to determine the coefficient β and even the coefficient α. Some of them only require knowledge of the direct solar total irradiance data, while the other need one or two spectral irradiance measurements [1]. They are formulated by Louche's et al. [31], Gueymard and Vignola [32], Pinazo et al. [33], Grenier et al. [34], Method U as stated by Casiniére et al. [1], Volz method [17], direct method and linear fitting, as cited by Cachorro et al. [35], spectral window method, as stated by Kaskaoutis et al. [36]. Junge [37] was the first to explore a relationship between the Ångström exponent α and the aerosol size distribution by assuming a power-law relationship for the number density (N) of aerosols as function of radius (r). Additionally, using Mie theory for spherical particles in the range 100 nm < r < 1000 nm, Junge was able to show that $\alpha \approx \nu - 2$ for nonabsorbing aerosols with $\alpha > 1$ as reported by Kaskaoutis et al. [38]. According to EI-Hussainy and Omran [39] Ångström [40] discussed a rather simple method to determine the true values of both Ångström parameters β_0 and α_0 based on the assumption that the parameter β is constant for all points of the spectrum provided it is computed by using the real value of α. Various studies [41–43] as reported by [37] have shown that the Ångström formula is a special case of a more complicated law valid for a limited range of values on the diameter of particles and for a limited interval of wavelengths. A method of determining the true turbidity coefficient β_0, corresponding to the real value of α, was proposed by Herovanu [44] and simplified by Ångström [45] as reported by Joseph and Manes [10]. This method is based on the assumption that β and α of Ångström's empirical aerosol extinction law can be determined by considering the intensities measured within two spectral ranges, 0–525 nm and 525–630 nm, with OG1 and RG2 cutoff filters. Additionally, Rodhe [46] has proposed a computational procedure for Ångström turbidity coefficient β. Volz [47] proposed it is possible to determine the values of β and α simultaneously from aerosol transmittances at two wavelengths using sun photometers as cited in Utrillas [48]. According to Casiniére et al. [1], Dogniaux [49] developed the relationship, as reported by Page [50] from a statistical analysis of the many β values determined in meteorological stations (by means of pyrheliometers fitted with standard color filters such as OG1, RG2, and RG8).

King and Byrne [51] showed that the size distribution of aerosols does not typically follow the Junge law nor does it take values from zero to unity as reported by Kaskaoutis et al. [52]. According to

Kaskaoutis et al. [37] King and Byrne [51] have shown that aerosol size distributions do not have radii extending from zero to infinity. Louche et al. [31] reported King and Buckius, based on the work of others, have given a simple expression relating visibility to α and β. With the help of the Bird and Hulstrom [53,54] parameterization of the direct solar total irradiance, Louche et al. [31] described a simple approach to determine Ångström turbidity coefficient β under cloudless skies. Melice et al. [55] calculated turbidity coefficients in the Sahara of Tozeur in southern Tunisia. They found high turbidity values in spring and summer. Since these turbidity indices differed slightly in their meaning, correlations and relationships were established between them as stated by EI-Hussainy and Omran [39].

Casiniére et al. [1] reported Nicholls [56] simultaneously derived the dependence of the parameter α on the wavelength. Bird [57] observed that turbidity versus wavelength often exhibits a curve-like relationship on a log–log plot. Different procedures can be used as stated by Cachorro [35] and Kaskaoutis et al. [52] for the purpose of determining Ångström parameters namely, Volz method, direct method, linear fitting and spectral window method. EI-Hussainy and Omran [39] reported Louche et al. [31] and Fox [58] estimated Ångström turbidity parameters using measurements of total spectrum direct irradiance. O'Neill and Royer [60] analyzed α computed from five VIS and NIR wavelengths (380–862 nm), and found a relationship between the effective radius in the aerosol size distribution and the value of α as cited by Kaskaoutis et al. [59]. As reported by Basart et al. [29], Kaufman [61] pointed out that negative values of the difference $\delta\alpha = \alpha(440{,}613) - \alpha(613{,}1003)$ indicate the dominance of fine mode aerosols, while positive differences reflect the effect of two separate particle modes.

Gueymard [62,63] expressed the turbidity as Bird [57] did, two different values for α: α_1 for $\lambda < 500$ nm and α_2 for $\lambda > 500$ nm as reported by Casiniére et al. [1]. As cited in Janjai [24] Kaufman et al. [64] argue the need to measure aerosol characteristics at different locations and in various meteorological conditions to fully assess regional aerosol properties. Grenier et al. [34] presented three linear equations valid for three ranges of columnar water vapor content w for calculating the Ångström turbidity coefficient β from T_L as reported by Kasten [65]. Pinazo et al. [33] propounded new method for calculating the Ångström turbidity coefficient based on the ratio of direct solar radiation to global solar radiation on a horizontal surface and on the "*C*" model of Iqbal. Utrillas [47] reported Bokoye et al. [66] have shown that the optimum wavelength pair in the 290–900 nm interval is 400 nm and 750 nm when using the Volz method. Gueymard [32] has proposed a semi-physical method to evaluate turbidity from broadband irradiance measurements and other atmospheric parameters. This method requires cloudless skies and very accurate measurements of global (or diffuse) and direct radiation. Eck et al. [67] showed how, in the wavelength range of 340–870 nm, α can increase by a factor of 3–5 as wavelength increases for biomass burning and urban aerosols, while remaining constant or decreasing in the presence of mineral dust.

Malik [68] modified Pinazo et al.'s method and developed two new correlations that link Pinazo's method with model '*C*' by Iqbal. The modified model was used to compute Ångström turbidity coefficient for Brunei Darussalam using measured data on solar radiation for a period of 5 years (1990–1994). Adeyewa and Balogun [69] reported Adeyefa et al. [70] also examined spectral and broad-band solar radiation measurements made non-continuously between 1990 and 1994 at Abisko. The study showed that there was an increase in atmospheric turbidity due to the huge amounts of volcanic aerosols injected into the stratosphere by the volcanic eruptions of Mount Pinatubo in June 1991. Utrillas [48] developed a new method aimed at determining the Ångström turbidity coefficients from measurements of broadband direct irradiance with filters. The results obtained using this method have been compared with those obtained using the Ångström method, and both have been compared with the values obtained from 1000-nm wavelength spectral irradiance measurements. O'Neill et al. [71] demonstrated that an Ångström exponent-based separation of coarse from fine mode contribution to τ is feasible in part because of the coarse mode τ spectral variation being approximately neutral. Schuster et al. [72] addressed the link between Ångström exponent curvature and the ratio between fine and total aerosol volume. Recently, Gobbi et al. [73] introduced a straight-forward graphical frame-work that allows to discriminate different aerosol types based on aerosol spectral measurements by sunphotometers which can be characterized by three independent pieces of information: τ, α and the spectral curvature of α ($\delta\alpha$). Plotting data in this space allows for inference of the aerosol fine mode size and fractional contribution to total τ.

Tables 1–4 describe summarizing studies which have been conducted at different regions by using various methods that are mentioned above to determine Ångström turbidity coefficients.

3. The Linke Turbidity coefficient (T_L)

The Linke turbidity factor T_L is an indicator of the number of clean dry atmospheres that would be necessary to produce the attenuation of the extraterrestrial radiation that is produced by the real atmosphere. Typical values range from 1 to 10 [28]. T_L can be obtained directly from observations performed during very clear sky periods, but this kind of experimental data is rarely available, thus T_L is generally an estimated parameter. Also, time series of radiation data are generally too short to allow estimation on a daily basis. Fortunately, long-term monthly average values, T_{Lm}, are sufficient for most applications [26]. Feussner and Dubois [14] published a series of spectral data tables enabling the calculation of $\delta_R(m)$ where both molecular scattering and absorption by the stratospheric ozone layer are taken into account. Additionally, Dogniaux [89] derived empirical relation from extensive experimental campaigns, illustrating the observed variation of Linke's turbidity factor with solar elevation angle, h (°), atmospheric water vapor content w (cm) and Ångström's turbidity coefficient β as reported by Molineaux et al. [90]. According to Kasten [65] in order to facilitate the determination of $T(m)$ from pyrheliometric measurements, Kasten [91] presented a simple analytic formula for $\delta_R(m)$, based on numerical values published by Feussner and Dubois [14] and internationally recommended by CSAGI [92]. Based on their new values of $\delta_R(m)$ Louche et al. proposed a new analytic formula for the function $\delta_R(m)$ and demonstrated that the new $\delta_R(m)$ values are definitely higher than the old CSAGI values used by Kasten, particularly at small relative optical air masses m.

Ineichen and Perez [93] reported that Kasten [91] fitted the following equation to those tables published by Feussner and Dubois [14]

$$\delta_{cda} = (9.4 + 0.9 \times AM)^{-1}$$

which is known as Kasten's pyrheliometric formula. Additionally, in an attempt to improve the formulation proposed by Kasten [91], Louche et al. [94] and Grenier et al. [34,95] added absorption by the permanent gaseous constituents to the definition of $\delta_R(m)$ (these gases are considered uniformly mixed and invariable in both a clean dry atmosphere and a turbid atmosphere). As cited in Navvab et al. [21], Valko [96] has suggested empirical fits of the Linke turbidity factor (T_L) as a function of aerosol and water vapor content and relative air mass. Grenier et al. [34] reported in 1986 that a determination of $\delta_R(m)$ based on more accurate values of spectral extraterrestrial solar irradiance and extinction coefficients of the various attenuators was carried out by Louche who proposed the following algorithm to evaluate the optical thickness of the CD

Table 1
Studies of atmospheric turbidity in Europe using Ångström formula.

Author	Year	Method	Area	Results and conclusion
Kaskaoutis et al. [59]	2006	Based on ground spectroradiometric measurements taken in the atmosphere of Athens during May 1995 Ångström exponent values, α, have been derived at five narrow spectral bands, 340–380, 380–440, 440–500, 500–670 and 670–870 nm using the Volz method as well as in the whole spectrum (340–870 nm) using both Volz and least-squares methods	Greece	1. α values depend strongly on the wavelength interval used for their determination. Very high α values (2.83 ± 0.66) were derived in the 340–380 nm band, while as the wavelength interval used for α determination shifts toward longer wavelengths, lower α values were computed, (α = 0.53 ± 0.39) in the 670–870 nm 2. The computed α values in the whole spectrum are positively correlated with the other α determined in the narrow spectral intervals 3. The two methods, Volz and least squares fit, result in the same α values in the 340–870 nm region
Kaskaoutis and Kambezidis [36]	2008	Used spectral direct-beam irradiance measurements at five narrow spectral intervals (i.e. 340–380 nm, 380–440 nm, 440–500 nm, 500–675 nm, 675–870 nm) as well as the whole spectrum, 320–1000 nm taken in Athens during May 1995, to investigate the ability of the different techniques mainly; the Volz method (VM), the direct method (DM) and the least-squares method (LM) to derive similar Ångström turbidity coefficients and their dependence on the spectral range used for their determination	Greece	1. The various techniques lead to different Ångström turbidity coefficients especially if narrow spectral bands at the shorter wavelengths are used 2. The Ångström turbidity coefficients derived by any of the three methods at short wavelengths are not representative of the whole spectrum 3. As regards the whole spectrum, both LM and DM estimated quite similar β and α values, with the standard errors of each fit to be almost the same 4. On the other hand, when narrow spectral intervals are used for their determination, the errors in β and α increase
Cucumo et al. [74]	1999	Carried out experimental determination and estimation of the Linke turbidity factor and Ångström turbidity coefficient using experimental data of beam solar radiation taken at Arcavacata di Rende, at the University of Calabria from August 1995 to August 1996, and at Casaccia (Rome) for the period May 1992–March 1995, and a method introduced recently by Pinazo et al. for β and Dogniaux's formula for T_L	Italy	1. The Dogniaux formula allowed the calculation of the Linke turbidity factor with average errors of about 20% and quadratic errors of about 30% assuming a suitable constant value of the Ångström coefficient 2. The Ångström turbidity coefficient can be calculated in a simple way, using the Katz formula, with an average error of about 10% and an average quadratic error of 30% if the experimental value of the Linke turbidity factor is inserted into the Katz formula
Wagner and Silva [75]	2008	Performed a simulation study in order to show the influence of the aerosol optical depth (τ) distribution together with the corresponding error distribution on the resulting Ångström exponent (AE) distribution	Portugal	1. When the τ frequency distribution is wide (e.g. sigma = 2) then the resulting Ångström exponent frequency distribution is narrower in comparison with a narrow AOD-distribution (e.g. sigma = 1.1) 2. If the mean τ is high (e.g. 0.4) the Ångström exponent frequency distribution is narrower as for low turbidity values (e.g. 0.06) in case of identical absolute errors 3. If the relative errors at both wavelengths are equal, or in a practical sense similar, then the peak or maximum of the Ångström exponent distribution reflects the true value, otherwise a shift either to smaller or to higher Ångström exponent values will occur 4. Non symmetrical errors result in a shift of the maximum of the Ångström exponent distribution 5. The Ångström exponent distribution is less symmetric (higher skewness) when the τ distribution is narrow and more symmetric for a wide τ distribution

Table 1 (*Continued*)

Author	Year	Method	Area	Results and conclusion
Kaskaoutis and Kambezidis [76]	2006	Compared the aerosol optical depths (τ) derived by the Ångström's formula by means of direct-beam spectra measurement obtained by a LICOR spectroradiometer in the 300–1100 nm wavelength range at several sites in the Athens basin under cloudless skies and different atmospheric conditions during May 1995 with the spectral aerosol optical depths (SAODs) retrieved through the SMARTS model and, therefore, to check the validity of the former From the available SAODs derived from SMARTS model, the AODs at two selected wavelengths (500 and 1000 nm) were used for further analysis	Greece	1. The Ångström's formula is capable of estimating SAOD values 2. The derivation of the τ values obtained from the estimation of α and β values was accurate as it correlated well with the SMARTS-estimated SAOD values 3. The correlation coefficient between SAOD and τ was always above 0.7 and in the majority of the cases its values were above 0.9 4. The correlation coefficient was found to depend strongly on the Ångström's turbidity parameter β and to increase as β increases 5. At 500 nm the τ estimation overestimated the derived SAODs from the model resulting in negative difference values 6. At 1000 nm, the respective differences were higher than those at 500 nm
Polo et al. [77]	2009	Determined the Ångström parameters and total ozone column obtained by means of a high-resolution spectroradiometer of selected clear sky days during one year for a characteristic semi-desertic site in the south-east of Spain	Spain	1. The turbidity and wavelength exponent were coherent with those reported by different authors for desertic area 2. A seasonal behavior of relative low turbidity values around 0.02–0.1 was identified during the autumn and winter days, and higher values in the range of 0.2–0.6 were found for spring and summer 3. Some events of high turbidity (β around 0.3–0.4) with low wavelength exponent (around 0.4–0.6 μm) have been observed eventually for several days in spring and summer 4. Good agreement between total ozone column values determined experimentally with the daily evolution reported by TOMS
Pedrós at el. [78]	1999	Determined and analyzed the Ångström turbidity coefficient, the Linke turbidity factor, and the Unsworth–Monteith coefficient based on measurements of normal direct irradiance and global horizontal irradiance taken in Valencia, Spain, between January 1990 and December 1996	Spain	1. The values of all the considered turbidity coefficients have a minimum in the winter and grow progressively, reaching a maximum in the summer 2. The summer values are at least twice as high as the winter ones 3. The morning turbidity values are higher than the afternoon ones, and all three turbidity coefficients vary in a similar manner throughout the day 4. The correlation between the coefficients is very good with correlation coefficients near to unity
Cañada et al. [79]	1993	Employed direct radiation measurements performed in Valencia to derive the monthly values of Ångström's turbidity coefficient, β, under cloudless skies by means of Louche et al. method and then compared with models reported for Avignon, Ajaccio and Dhahran	Spain	1. The turbidity coefficient of Valencia showed higher and lower values of β than Avignon and Ajaccio in summer and winter respectively. This is mainly due to the effect of the air mass origin, the air temperature and the water vapor content 2. The variation in the monthly average values at Valencia was of a similar trend to that of Avignon, Ajaccio and Dhahran
Cachorro et al. [80]	2001	Used two methods namely, the direct method and the window method and spectroradiometric direct irradiance measurements at four non-absorbing spectral windows: A (370–490 nm), B (749–754 nm), C (776–782 nm), and D (862–872 nm) under clear sky conditions at two different climate stations in Spain, in a study of the variation in the Ångström turbidity parameter α and its dependence on the spectral range used in its determination	Spain	1. The α-values determined in the UV were very different than those determined in the VIS and VIS–NIR ranges 2. The low values determined in the VIS–NIR interval contrast with the high values given by the 350–400 nm interval 3. The differences between experimental and estimated values were as: • For the 340–400 nm range was 1% (s.d. = 7%) and −2% (s.d. = 22%) for the 350–400 nm range, respectively • For the VIS and VIS–NIR spectral ranges gave similar results • The VIS range had a mean value of 26% (s.d. = 15%) and the VIS–NIR is 32% (s.d. = 16%) 4. These results seem to indicate that the VIS and VIS–NIR ranges are not useful predictors of the AOD in the UV.

Table 2
Studies of atmospheric turbidity in Asia using Ångström formula.

Author	Year	Method	Area	Results and conclusion
Janjai et al. [8]	2003	Determined Ångström's turbidity coefficient, β, for 53 meteorological stations covering Thailand by using three different methods mainly: (a) a calculation of β based on narrow-band spectral irradiance; (b) a derivation of β from broad-band direct irradiance; and (c) an estimation of β from visibilities	Thailand	1. For most stations in the north, the northeast and the central region, values of β were relatively high in dry season (November–April) and low in the wet season (May–October) 2. For the dry season, it was inferred that the northeast monsoon together with the high convection in the summer was responsible for the higher values of β for these regions 3. For the lower values of β in the wet season, it is due to the effects of rains which remove aerosols from the atmosphere 4. For the south, values of β are relatively low, and remain nearly constant for the whole year. It is presumably caused by rains from the two monsoons, which constantly wash through the atmosphere almost all year round
Hussain et al. [7]	2000	Determined atmospheric turbidity parameters following Ångström's method with the help of a pyrheliometer and cut-off glass filters for several months of 1990, 1991, 1995 and 1997 for Dhaka (latitude 23.7°N) and for the months of March and April 1990 for two rural locations, Sripur (24.11°N) and Haripur (26.03°N), respectively	Bangladesh	1. Considerable variation of Ångström's turbidity parameters for Dhaka over the year was observed with a maximum value in March 2. The value for Sripur for the month of March was somewhat lower 3. The value for Haripur for the month of April was not much different from that of Dhaka
Janjai s. et al. [24]	2009	Employed Bouguer's law to examine aerosol optical depth and Ångström coefficients for three sites in Bangkok and suburbs under cloudless conditions Three types of Sun photometers have been used to measure direct normal spectral irradiance in six wave-length bands at a frequency of 1 min at these sites for a period of 2 years (2004–2005); Yankee MFRSR for Nakhon Pathom (NP) at wavelengths 413, 500, 613, 671, 863 and 940 nm, EKO MS-110 for Phatum Thani (AIT) at wavelengths 368, 500, 675, 778 and 863 nm and CIMEL CE 317 for Bangkok (BK) at wavelengths 380, 440, 500, 670, 870 and 937 nm.	Thailand	1. Aerosol optical depth showed a marked change for all bands. They were lowest in October/November and gradually increase during the dry season, reaching maximum values in late March and early April 2. The Ångström exponent α also undergoes a seasonal variation, with maximum values in the dry season and minimum values at the height of the wet season 3. There was much more scatter in the seasonal pattern of the turbidity coefficient β, with maximum and minimum peaks occurring at various times of the year
Li and Lam [28]	2002	Used three methods namely, Louche's model and Pinazo's model and visibility method to determine Ångström's turbidity coefficient (i.e. β_{Lou}, β_{Pin} and β_{Vis}) based on the 9-year (1991–1999) horizontal global and diffuse solar irradiances measured at the City University of Hong Kong. With the three sets of Ångström's turbidity data, the Linke turbidity factors (i.e. T_{Lou}, T_{Pin} and T_{Vis}) and the original Linke's expression (T_{Lin}) have also been computed using Dogniaux's formula. In addition, the Linke turbidity also determined based on the original Linke's expression (T_{Lin})	Hong Kong	1. The annual averages of the Ångström turbidity coefficients based on the Louche, Pinazo and visibility methods were 0.132, 0.165 and 0.156, respectively 2. In general, β_{Pin} produces the highest monthly average values followed by β_{Vis} and then β_{Lou} 3. The T_{Lin} has the lowest monthly average values among the four models, ranging from 3.7 to 5.26 4. For the other three models, the minimum monthly average occurs at T_{Lou} in December at 4.01 while the monthly peak appears at T_{Pin} in May at 6.72 5. It is envisaged that in Hong Kong the clear sky conditions can be defined as between turbid and clear
Ganesh et al. [81]	2011	Ångström's turbidity parameters, α and β are calculated and analysis on daily, monthly, seasonal and annual using the solar radiation data collected at five optical channels at a continental station Mysore, India during December 2003–June 2006 performed by a portable sunphtometer MICROTOPS II	India	1. The observations show that α and β vary throughout on individual day because of changes in the atmospheric meteorological parameters 2. It is observed that β is highest during summer and lowest during winter 3. An anti-correlation between α and β is observed throughout the day during all seasons indicating continuous redistribution of fine and coarse particles under the influence of meteorological parameters
Balakrishnalah et al. [82]	2011	Used Multi-Wavelength solar Radiometer (MWR) measurements preformed during 2007–2008 over tropical semiarid site of Anantapur, India for determining the temporal variability of aerosol characteristics	India	1. The mean and standard values of the Ångström exponent, α (turbidity coefficient, β), are found to be 1.06 ± 0.33 (0.17 ± 0.05) during winter, 0.97±0.35 (0.22 ± 0.04) during summer and 0.85 ± 0.26 (0.16 ± 0.06) during monsoon 2. The largest mean monthly AOD in February to April are associated with large a values indicative of fine-mode particles

Table 3
Studies of atmospheric turbidity in America using Ångström formula.

Author	Year	Method	Area	Results and conclusion
Pinker et al. [83]	2004	Used an automatic sun-tracking sky radiometer of the CIMEL type (Model CE 318) to measure both sun and sky radiance at eight spectral channels. And then used Dubovik and King method (2000) to calculate the volume size distribution of τ from the direct solar and diffuse sky radiance measurements, water vapor absorption band (940 nm) to retrieve precipitable water vapor and the Ångström exponent to calculate aerosol optical thickness	United States	1. Annual means of spectral τ were found to be low over this semiarid region and they agree well with historic data collected two decades ago at a nearby location 2. Significant month-to-month variations in volume size distribution, and, in particular, considerable variation in accumulation mode particles was evident 3. Monthly means of the Ångström exponent were found to be higher during summer, suggesting an abundance of submicron-sized particles 4. Precipitable water vapor retrieved from the 940-nm channel showed a minimum of ~0.4 cm in February and a maximum of 2.5 cm in August. These values were compared with NOAA/NCEP Eta model forecast values and found to be in good agreement
Sapkota and Dhaubhadel [84]	2002	Utilized a Volz multispectral sunphotometer, equipped with interference filters of 3% bandwidth centered at 440, 500, 640 and 880 nm wavelengths for the determination of the atmospheric turbidity over Kathmandu Valley during the pre-monsoon period during the pre-monsoon period of 1999 The Meteorological Research Inc. (MRI) integrating nephelometer model 1550 was also operated to measure the air quality of the valley The nephelometer and sunphotometer were operated at a height of about 45 ft from the ground level	Nepal	1. About 76.8% of the observed Ångström coefficient β lied above 0.2 and about 14.28% lied above 0.4 2. About 86.9% of the observed scattering coefficient b_{scat} lied above 0.210 km^{-1} which indicates heavy pollution 3. The classes of maximum occurrence for β and b_{scat} were found to be 0.22–0.28 and 0.20–0.30 km^{-1}, respectively 4. The valley is at the verge of extremely heavy air pollution
Fox [58]	1994	Used the method of Louche et al. [31] to determine Ångström's turbidity coefficient, β, from measurements of direct normal irradiance (broadband) collected by the University of Alaska Geophysical Institute from 1 June 1979 to 30 June 1983	Alaska	1. The method was sensitive enough to detect a seasonal pattern along with annual deviations associated with hemispheric scale disturbance created by the El Chichón volcanic eruptions of 1982 2. Knowledge of the magnitude and variability of Ångström's turbidity coefficient will prove useful in modeling solar radiation input to northern ecosystems
Salazar [85]	2011	Used the mean monthly global solar irradiation values measured at ten sites in Argentina from the former Argentinean radiometric network REDSOL to estimate the monthly average Ångström turbidity coefficient β values	Argentina	The monthly average value estimates of turbidity coefficient β show expected behaviour, with values varying according to the season of the year

(clean dry) atmosphere, from the relative air mass. Abdelrahman et al. [97] calculated the Linke factor and the Ångström turbidity coefficient β at Dhahran (Saudi Arabia). They found that their values are larger in summer than in winter. They constructed a linear relation between the two parameters. Their coefficients were quite different from those calculated at Avigon (France) and Potsdam (West Germany). They attributed these differences to the desert climate at Dhahran as stated by El-Hussainy and Omran [39].

Grenier et al. [34] repeated the calculation of new $\delta_R(m)$ values and proposed a fourth order polynomial for the function $\delta_R(m)$ similar to Louche et al. but with slightly different coefficients, because they confined the range of validity to $m \leq 6$ in order to achieve a better fit in this range. Further, Grenier et al. separated the influence of the Ångström turbidity coefficient β and of the columnar water vapor content w of the atmosphere on the Linke turbidity factor $T_L(m)$ and displayed the variation of $T_L(m)$ with relative optical air mass m for various values of β and w as parameters. Grenier et al. derived algorithms for converting $T_L(m)$ into a standardized Linke turbidity factor T_L at relative optical air mass $m = 2$ with the help of polynomials of the third order in $T_L(m)$ whose coefficients are polynomials of the fourth order in m as reported by Kasten [65]. Molineaux et al. [90] reported that Grenier et al. [34] in the same approach, added some minor changes to the spectral absorption and scattering equations and obtained the polynomial, yielding very similar values to Louche's [94] relation. As reported by Grenier et al. [95] recently, a model of Linke's turbidity factor was developed by the authors (Grenier et al. [34]), in terms of spectral extraterrestrial solar irradiations and extinction coefficients of atmospheric absorbers. With the help of the model, the authors calculated the theoretical variations of T_L from the values of the two atmospheric

Table 4
Studies of atmospheric turbidity in Africa using Ångström formula.

Author	Year	Method	Area	Results and conclusion
Tadros et al. [86]	2002	Estimated the ÅTC, α and β from the τ using spectral broadband data. The data were obtained from pyrheliometric measurements in the period 1991–96 for two sites of different climatological and environmental view, Cairo in low Egypt and Aswan in Upper Egypt A technique used was similar to that of Volz (with two large pairs of spectral broadbands (bands 1/3 and bands 2/4) instead of narrow bandwidth), was applied to SMARTS2 and SPECTRAL2 models to select the suitable spectral broadband for estimating ÅTC	Egypt	1. The turbidity β started to increase from March, and peak in May with value 0.317 for Cairo and 0.179 for Aswan. This was attributed to Khamsin depressions coming from Great Sahara associated with advection of desert air masses with an extremely high concentration of aerosols 2. The maximum of β at summer months occurred on June (0.32) and August (0.178) at Cairo and Aswan, respectively. These maxima were due to hot and dry weather, wind dispersion of aerosols and dust raised from ground 3. The Sudan monsoon trough prevailing during the autumn months cause maximum 0.328 at September in Cairo and 0.156 at October in Aswan 4. The atmospheric turbidity depends on the local weather conditions, and the prevailing winds may transport aerosol particles from distant sources depending on the speed of wind and its direction. The increase in turbidity during 1992 is attributed to the eruption of Mount Pinatubo in the Philippines on 15 June 1991
El-Hussainy and Omran [39]	1998	Calculated and analyzed the Linke turbidity factor, aerosol optical thickness and the Ångström turbidity parameters using measurements of both total spectrum and broad spectral bands of direct irradiance. Relations were also constructed to estimate both aerosol optical thickness and the Ångström turbidity coefficient from Linke turbidity factor	Egypt	1. The annual variation showed lowest value of turbidity in winter and highest values in both spring due to the Khamsin disturbances with sandstorms, and summer due to hot air mass and large water vapor contents 2. Aerosol extinction coefficients decrease with the increase of both wavelength and optical air mass 3. Aerosol optical thickness could be estimated from the Linke turbidity factor values with errors below 20%. Aerosol concentration parameter β can be estimated with errors above 50%
Maduekwe and Chendo [87]	1997	Studied the trend in the atmospheric turbidity in Lagos, a tropical city in Nigeria, for 17 months between 1990 and 1991 with a Volz sunphotometer using the Ångström turbidity coefficient, β, at 500 nm wavelength	Nigeria	1. The variation of β indicated that the atmosphere was mostly turbid throughout the year 2. The aerosols influencing the city were both of maritime and Saharan origin coupled with locally produced particulates from industries and automobile combustion 3. The month of January experienced the highest turbidity, with a mean value of 0.497, while November experienced the lowest aerosol loading on average with the value of 0.225
Trabelsi and Masmoudi [88]	2011	Used the direct measurements taken by pyrheliometer during a year: July (2008) to June (2009) to derive Angstrom coefficient (β)	Tunisia	1. The level of turbidity found depends greatly upon the emissions and climate system of each area. The results show the highest turbidity in the summer season and the low turbidity in the winter season for each index 2. Values of β are higher in the coastal region than those in the island for all seasons, except in winter, β is almost the same in both areas. The low values of β in the island of Kerkennah can be explained by low amount of continental aerosols because in this island we have the major party of the territory is occupied by lagoons (Sebkha) 3. The turbidity of the atmosphere has a diurnal variation; it increases in the afternoon. This situation can be explained by a supplementary load of the atmosphere in water vapor due to a sea breeze circulation and the transport of aerosols from local transport into the site, essentially during summer afternoons

constituents at the origin of the turbidity, the water vapor, represented by its content w, and amount of aerosol, represented by Ångström's turbidity coefficient β and from the relative optical air mass m. Using the results of the model, they proposed to introduce a new quantity strictly representative of the atmospheric turbidity, the standardized Linke's turbidity factor, that is, the usual factor T_L from the dependence on m has been removed. Diabaté et al. [26] reported that ESRA [98] discusssed an approach for calculating T_{Lm}, and it is found that the analytical inversion of the clear-sky model is not required. So, it is easier to be implemented and operated than that of Cucumo et al. [99]. Ineichen and Perez [93] reported Molineaux et al. [90] noted that Louche and Grenier's expressions for $\delta_R(m)$ become divergent, respectively for air mass greater than 20 and 7. They adapted the coefficients of Linke's original expression to take into account the absorption by the permanent gases. Cucumo et al. [99] propose a method for calculating T_{Lm} from

Table 5
Number of measurements and studies of atmospheric turbidity in different areas in the world using Linke turbidity factor.

Author	Year	Method used	Area	Results
Mavromatakis and Franghiadakis [102]	2007	Used more than a year's data to determine the Linke turbidity coefficient through measurements of the global and diffuse irradiance at 5 min intervals with two CM11 pyranometers by Kipp Zonen at an altitude of 122 m	Greece	1. Typical values of the Linke turbidity lied in the range of 2.3–3.5, while the minimum and maximum measured values were 2.0 and 4.0, respectively 2. Strong seasonal effects on the monthly averaged T_L values were not observed, while their profile displayed a smooth transition from lower to higher values
Chaâbane et al. [25]	2004	Used pyrheliometric measurements in a coastal tourist location in Tunisia (Sidi Bou Saïd), during three summer months (June, July and August 1999) to compute the Linke turbidity factor	Tunisia	1. The increase of T_L is due to industrial, increase of traffic in the site and transport of both moisture from maritime air masses and Saharan dust carried by sirocco winds 2. The atmospheric turbidity depends on the local weather conditions 3. Prevailing winds may transport aerosol particles from various sources depending on the speed of the wind and its direction
Chaiwiwatworakul, Chirarattananon [103]	2004	Derived the turbidity indices, namely, Linke factor, Ångström coefficient, and illuminance turbidity factor from the record of two and half-year measurement derived directly from measurements taken by pyrheliometer, Volz sunphotometer and beam illuminance meter at the Asian Institute of Technology (AIT), to investigate atmospheric turbidity for tropical Thai sky	Thailand	1. Values of the turbidity of the atmosphere are found to vary with seasons, months and with time of day 2. Atmospheric turbidity is low and quite stable during dryer months and increases in wet season from March to August 3. The turbidity increases from morning to noontime and then decreases in the afternoon 4. The annual mean values of Ångström coefficient, wavelength exponent, Linke factor and illuminance turbidity factor are 0.098, 1.272, 3.306 and 2.818, respectively
Diabaté et al. [26]	2003	Used Aguiar method 1995, for calculating T_L from a time series of daily irradiation	9 stations in Egypt, 2 in Mozambique and Zimbabwe, 1 in Algeria, Tunisia and Zambia	1. T_L is almost constant throughout the year close to the Mediterranean basin with values around 3.5 2. Stations located in the sub-tropical Southeastern part exhibit large variations of T_L The range and the behavior of the monthly average values of T_{Lm} throughout the year, demonstrate that T_{Lm} is related to the climate, an unsurprising result
Rapti [104]	2000	Used global solar and diffuse sky radiation data, measured with a Moll-Gorczynski pyranometer for three values of relative atmospheric mass which correspond to the solar elevations of 20°, 28° and 40° in a coastal middle latitude Mediterranean location, for the time period 1973–1976 measured in Greece to investigate the diurnal and seasonal variations of the atmospheric turbidity using Linke's factor	Greece	1. Atmospheric transparency decreased with decreasing relative optical air mass. This leads to a virtual increase of Linke's turbidity factor with increasing solar elevation 2. Diurnal and seasonal variations of atmospheric transparency were found with a summer afternoon minimum and a winter morning maximum 3. Real diurnal and seasonal variations of atmospheric turbidity were found with a winter morning minimum and a summer afternoon maximum 4. The largest diurnal turbidity variation is observed in October (about 6%). The annual variation of turbidity is estimated to be about 40%, between the morning value of turbidity in February and the afternoon value in June

Table 5 (*Continued*)

Author	Year	Method used	Area	Results
López and Batlles [9]	2004	Compared and evaluated the performance of three turbidity algorithms – Dogniaux's algorithm, Louche's algorithm and Gueymard's algorithm – based on broadband solar irradiance measurements in the estimation of β The evaluation of the performance of the models was undertaken by graphical and statistical (root mean square errors and mean bias errors) means	Almería, Bondville, Desert Rock, Fort Peck, Goodwin Creek, Granada, Penn State, Table Mountain	1. Dogniaux's and Louche's algorithms provide hourly turbidity estimates closer to each other than by using Gueymard's method 2. Differences on turbidity estimates by Dogniaux's and Louche's algorithms were notably reduced by introducing a correction term depending on precipitable water 3. Ueymard's method is the more accurate and reliable, if information about the wavelength exponent is not available 4. Turbidity algorithms by Dogniaux [86] and Louche et al. [31] have shown to be less sensitive to broadband solar radiation data affected by cloud interference than that by Gueymard and Vignola [32]
Coste and Eftimie [105]	2010	Use (Kasten and Young Model, and Remund and Page Model) to calculate the Rayleigh optical thickness (δ_r) and the relative optical air mass (m) and then to determine the Linke turbidity factor	Romania	1. The turbidity factor had a decreasing variation during a day 2. At the sunset time, the minimum values of the turbidity factor were obtained 3. During the morning, the turbidity factor has higher values
Kryza et al. [106]	2010	Use the formula proposed by Dogniaux (1984) to determine the Linke turbidity factor	Poland	The formula proposed by Dogniaux (1984) was found to be in good agreement with measurements by Jacovides (1997)

a time series of daily irradiation. This method is based on (i) the computation of the direct and diffuse components by the means of the Collares–Pereira and Rabl correlation, (ii) the modeling of each component as an analytical function of T_{Lm} and (iii) the inversion of these models. The method was applied to several sites in Italy. For these sites, the authors used the maximum daily irradiation values read per month in the ESRA [100]. These maxima are those found in the time-series of daily irradiation spanning the period 1966–1975. Additionally, Walkenhorst et al. [101] processed time-series of hourly means of beam irradiance. The Linke turbidity factor was calculated for the 4 h around noon for each day by inverting the ESRA formula for the clear sky beam irradiance. From the resulting calculated set of T_L for each month, the smallest three are averaged and this mean serves as an approximation for T_{Lm} as reported by

Table 6
Comparison between the Linke turbidity factor and Ångström turbidity coefficients.

Linke turbidity factor T_L	The Ångström turbidity coefficients, β and α
1. Useful parameter for comparison of cloudless atmospheric conditions	1. Ångström's turbidity coefficient is calculated from spectral direct irradiance at two wavelengths, usually 380 and 500 nm, in a part of the solar spectrum where absorption is negligible
2. Refers to the whole spectrum, that is, overall spectrally integrated attenuation, which includes presence of gaseous water vapor and aerosols	2. Ångström's turbidity coefficient β is usually determined under the assumption of a constant wavelength exponent $\alpha = 1.3$
3. It is a key input to several models assessing the downwelling irradiance under clear skies that are used by several communities in the fields of renewable energies, climatology, agro-meteorology, and atmospheric pollution	3. The expression does not permit separate estimation of the scattering and absorption optical depths. This causes important errors in the estimation of the diffuse solar radiation
4. Linke's turbidity factor has been less used than Ångström coefficient, although its experimental determination seems to be easier	4. The determination of Ångström coefficients is limited to small optical depth values. Therefore large optical depths that occur for large zenith angles cannot be considered
5. Does not exclusively explain the turbidity caused by aerosol because the quantity is also affected by the absorption of solar radiation in the visible and near IR regions by water vapor	5. The value of β can be affected by a change in the amount of atmospheric water vapor
6. Is not a pure turbidity coefficient. It also depends on air mass, m, and precipitable water, w even when the atmospheric conditions remain constant.	6. The Ångström formula is only a convenient approximation. It is not necessarily valid over an extended spectral range, and Ångström parameters that work properly in a given range might not be so appropriate in another. Nevertheless, the Ångström formulas seem to provide a good parameterization of spectral aerosol optical thickness for the 400–670 nm band
7. Intricate relation combining aerosol and water vapor transmittances integrated over the solar spectrum	7. The Ångström method suffers from the same problem that the methods based on photometer measurements employing the Volz technique do, that is, it is highly dependent on the selection of the spectral bands because different band combinations produce different α and β values
8. Difficult to determine accurately and fails to distinguish the effects of water vapor from the absorption due to aerosols	8. The major drawback to Ångström's formalism lies in the difficulty in getting a precise evaluation of α and β

Diabaté et al. [26]. Ineichen and Perez [93] developed a new formulation for the Linke turbidity factor (T_{LIP}) with the objective of removing its dependence upon solar geometry and fully compatible with the original formulation at air mass 2. They also developed two new simple clear sky models for global and direct normal irradiance.

Table 5 summarizes a number of measurements and studies of atmospheric turbidity in different areas in the world using Linke turbidity factor.

4. Comparison between Ångström's turbidity parameters and Linke turbidity coefficient

Table 6 shows the comparison between Linke turbidity factor T_L and Ångström turbidity coefficients, β and α.

5. Conclusions

The results of this overview indicate:

1. The atmospheric turbidity depends on local weather conditions and on the climate of the site. The prevailing winds, which may transport moisture or aerosol particles from distant sources, play a major role on the seasonal variation of turbidity.
2. During the last century, a number of atmospheric turbidity indices have been introduced and several methods were developed to determine their values. The most currently used are the Ångström turbidity coefficient β, the Linke turbidity factor T_L, Schuepp's turbidity coefficient B and Unsworth–Monteith TU.
3. Linke's turbidity factor T_L represents turbidity caused by aerosol and by water vapor. T_L is usually deduced from measurements of the direct component of solar radiation using pyrheliometric methods during very clear sky periods, but this kind of experimental data is rarely available, thus T_L is generally an estimated parameter.
4. In a pure Rayleigh atmosphere the turbidity factor is equal to one ($T_L = 1$). The closest value to this ideal value is achieved in extremely clear, cold air at high latitudes ($T_L = 2$). However, for a heavy polluted atmosphere the turbidity factor can increase to 8. Monthly mean values of T_L are normally about 3 for temperate rural climate, about 4 between the tropics and 5 for humid polluted atmospheres. The value for Linke turbidity factor specifically recommended by the branch literature for Europe is 3. But, at a local level the value of Linke turbidity factor can vary with the geographical coordinates, meteorological parameters and climatic conditions.
5. The Dogniaux formula allows the calculation of the Linke turbidity factor with average errors of about 20% and quadratic errors of about 30% assuming a suitable constant value of the Ångström coefficient. As suggested by Grenier [34,35] Linke's turbidity factor $T_{2.00}$ values, which correspond to the solar elevation of 30°, can be assumed to be independent of relative atmospheric mass and strictly representative of atmospheric turbidity.
6. The main problem with T_L is that it is difficult to be determined accurately and fails to distinguish the effect of water vapor from the absorption due to aerosol. Also it is not truly independent of the optical air mass and varies diurnally even with an unchanging atmosphere. Moreover, it is subject to virtual daily variations like broadband aerosol optical depth δ_a due to the combination of different parasitic effects. Virtual variations of T_L due to solar zenith angle, Z, alone are smoother than those of δ_a for $Z < 80°$ but become steeper for larger zenith angles. To remove a part of T_L's parasitic variation, it is advisable to calculate it for a reference air mass value of 2.
7. The Ångström exponent α is very popular because of the simplicity of the respective equation, because it enables to interpolate or to extrapolate aerosol optical properties. The Ångström turbidity coefficient β has long been used as a turbidity index and can be determined by different methods from spectral and broadband radiation measurements in the large spectrum band not affected by the water vapor absorption ($280 < \lambda > 630$ nm).
8. β cannot be evaluated unless α is known a priori. This uncertainty in α makes the determination of β less precise and discriminating. β can be calculated in a simple way, using the Katz formula, with an average error of about 10% and an average quadratic error of 30% if the experimental value of the Linke turbidity factor is inserted into the Katz formula.
9. Unlike the Linke turbidity factor, Ångström's turbidity coefficients may be employed in calculations of spectral direct and diffuse solar irradiance.

References

[1] Casiniére ADL, Bokoye AI, Cabot T. Direct solar spectral irradiance measurements and updated simple transmittance models. American Meteorological Society 1997;36:509–20.
[2] Wen CS. The fundamentals of aerosol dynamics. World Scientific Publishing Co. Pte. Ltd.; 1996.
[3] Iorga G, Stefan S. Effects of the atmospheric aerosol on the optical properties of cloud. Romanian Reports in Physics 2005;57(3):426–35.
[4] Stefan S, Iorga G, Zoran M. The atmospheric aerosols and their effects on cloud albedo and radiative forcing. In: Proceedings of the 2nd Environmental Physics Conference. 2006.
[5] Peixoto JP, Oort AH. Physics of climate. New York: Springer-Verlag, Inc.; 1992.
[6] Igbal M. An introduction to solar radiation. New York: Academic Press Canada; 1983.
[7] Hussain M, Khatun S, Rasul MG. Determination of atmospheric turbidity in Bangladesh. Renewable Energy 2000;20:325–32.
[8] Janjai S, Kumharn W, Laksanaboonsong J. Determination of Angstrom's turbidity coefficient over Thailand. Renewable Energy 2003;28:1685–700.
[9] López G, Batlles FJ. Estimate of the atmospheric turbidity from three broadband solar radiation algorithms. A comparative study. Annales Geophysicae 2004;22:2657–68.
[10] Joseph JH, Manes A. Secular and seasonal variations of atmospheric turbidity at Jerusalem. Journal of Applied Meteorology 1971;10:453–62.
[11] Linke F. Transmissions-Koeffizient und Trübungsfaktor. Beiträge zur Physik der Atmosphäre 1922;10:91–103.
[12] Kastrov VG. On the basic actinometric formula. Meteorological Bulletin 1928:7.
[13] Ångström A. On the atmospheric transmission of sun radiation and on dust in the air. Geografiska Annaler 1929;2:156–66.
[14] Feussner K, Dubois P. Trübungsfaktor. Beiträge zur Geophysik 1930;27:132–75.
[15] Schneider A. Trübungsfaktor and Trübungsdiagramm. Beiträge zur Geophysik 1937;51:69–99.
[16] Schuepp W. Die Bestimmung der atmospharischen Trubung ans aktinoroeter messungen. Archives for Meteorology, Geophysics, and Bioclimatology Series B 1949:257–346.
[17] Volz F. Photometer mit Selen-Photoelement zur spektralen Messung der Sonnenstrahlung und zur Bestimmung derWellenlängenabhängigkeit der Dunsttrübung. Archives for Meteorology, Geophysics, and Bioclimatology 1959;B10:100–31.
[18] Valko P. Vereinfachtes Auswerteverfahren für die Achüeppsch Methode zur Bestimmung der Atmöpharischen Trüburg. Archives for Meteorology, Geophysics, and Bioclimatology 1961;11:75–107.
[19] Unsworth MH, Monteith JI. Aerosol and solar radiation in Britain. Quarterly Journal of the Royal Meteorological Society 1972;98:778–97.
[20] Davis JA, Hai JI. Calculation of the solar radiation incident on a horizontal surface. In: Hay JE, Won TK, editors. Proceedings of the first Canadian solar radiation data workshop. Downsview: Canadian Atmospheric Environment Service; 1980. p. 32–58.
[21] Navvab M, Karayel M, Néeman E, Selkowitz S. Analysis of atmospheric turbidity for daylight calculations. Energy and Buildings 1984;6:293–303.
[22] Wen CC, Yeh HH. Analysis of atmospheric turbidity levels at Taichung Harbor near the Taiwan Strait. Atmospheric Research 2009;94:168–77.
[23] Wagner F, Silva AM. Some considerations about Ångström exponent distributions. Atmospheric Chemistry and Physics Discussions 2007;7:12781–805.
[24] Janjai S, Suntaropas S, Nunez M. Investigation of aerosol optical properties in Bangkok and suburbs. Theoretical and Applied Climatology 2009;96:221–33.
[25] Chaâbane M, Masmoudi M, Medhioub K. Determination of Linke turbidity factor from solar radiation measurement in northern Tunisia. Renewable Energy 2004;29:2065–76.

[26] Diabaté L, Remund J, Wald L. Linke turbidity factors for several sites in Africa. Solar Energy 2003;75:111–9.
[27] Jiakui T, Yong X, Tong Y, Yanning G, Guoyin C, Yincui H. Aerosol retrieval over land by exploiting the synergy of TERRA and AQUA MODIS DATA. Science in China Series D: Earth Sciences 2006;49(6):641–9.
[28] Li DHW, Lam JC. A study of atmospheric turbidity for Hong Kong. Renewable Energy 2002;25:1–13.
[29] Basart S, Pérez C, Cuevas E, Baldasano JM, Gobbi GP. Aerosol characterization in Northern Africa, Northeastern Atlantic, Mediterranean Basin and Middle East from direct-sun AERONET observations. Atmospheric Chemistry and Physics 2009;9:8265–82.
[30] Mächler MA. Parameterization of solar irradiation under clear skies. Thesis, Mechanical Engineering Department, University of British Columbia; 1983.
[31] Louche A, Maure M, Simonno G, Peri G, Iqbal M. Determination of Angstrom's turbidity coefficient from direct total solar irradiance measurements. Solar Energy 1987;38(2):89–96.
[32] Gueymard C, Vignola F. Determination of atmospheric turbidity from the diffuse-beam broadband irradiance ratio. Solar Energy 1998;63(3): 135–46.
[33] Pinazo JM, Cañada J, Boscá JV. A new method to determine Ångström's turbidity coefficient: its application for Valencia. Solar Energy 1995;54(4):219–26.
[34] Grenier JC, casinitére ADL, Cabot T. A spectral model of Linke's turbidity factor and its experimental implications. Solar Energy 1994;52(4):303–13.
[35] Cachorro VE, Frutos AM, Casanova JL. Determination of the Angstrom turbidity parameters. Applied Optics 1987;26(15):3069–76.
[36] Kaskaoutis DG, Kambezidis HD. Comparison of the Ångström parameters retrieval in different spectral ranges with the use of different techniques. Meteorology and Atmospheric Physics 2008;99:233–46.
[37] Junge CE. The size distribution and aging of natural aerosols as determined from electrical and optical measurements in the atmosphere. Journal of Meteorology 1955;12:13–25.
[38] Kaskaoutis DG, Kambezidis HD, Hatzianastassiou N, Kosmopoulos PG, Badarinath KVS. Aerosol climatology: dependence of the Angstrom exponent on wavelength over four AERONET sites. Atmospheric Chemistry and Physics Discussions 2007;7:7347–97.
[39] El-Hussainy FM, Omran MA. Analysis and trends of atmospheric turbidity parameters over Cairo. Meteorology and Atmospheric Physics 1998;66:113–21.
[40] Ångström A. Techniques of determining the turbidity of the atmosphere. Tellus 1961;8:214–23.
[41] Tomasi C, Caroli E, Vitale V. Study of the relationship between Ångström's wavelength exponent and Junge particle size distribution exponent. Journal of Climate and Applied Meteorology 1983;22:1707–16.
[42] Dellago C, Horvath H. On the accuracy of the size distribution information obtained from light extinction and scattering measurements. I. Basic considerations and models. Journal of Aerosol Science 1993;24:129–41.
[43] Cachorro VE, de Frutos AM. A revised study of the validity of the general Junge relationship at solar wavelengths: application to vertical atmospheric aerosol layer studies. Atmospheric Research 1995;39:113–26.
[44] Herovanu AM. Determination des parameters D' Ångström par des observations actinometriques courantes. Geofis Pura Appl 1959;44:315–21.
[45] Ångström A. The parameters of atmospheric turbidity. Tellus 1963;16:64–75.
[46] Rodhe B. Simplified formulae for computation of the turbidity coefficient. Sveriges Meteor. Hydrol. Inst. Serie, Meteor.; 1969. p. 20.
[47] Volz FE. Economical multispectral sun photometer for measurements of aerosol extinction from 0.44 nm to 1.6 nm and precipitable water. Applied Optics 1974;13:1732–3.
[48] Utrillas MP, Pedrós R, Martinez-Lozano JA, Tena F. A new method for determining the Ångström turbidity coefficient from broadband filter measurements. American Meteorological Society 2000;39:863–74.
[49] Dogniaux R. Variations géographiques et climatiques des expositions énergétiques solaires sur des surfaces réceptrices horizontales et verticales. Tech. Rep. Misc. B38. Institut Royal Mé-téorologique, Uccle, Belgium; 1975. Available from Kon. Meteo Institute, Ringlaan 3, 1180 Brussels, Belgium.
[50] Page JK. Prediction of solar radiation on inclined surfaces. Solar energy R&D in the European community series F, Solar radiation data, vol. 3. D. Reidel; 1986. p. 459.
[51] King MD, Byrne DM. A method for inferring total ozone content from spectral variation of total optical depth obtained with a solar radiometer. Journal of Atmospheric Science 1976;33:2242–51.
[52] Kaskaoutis DG, Kambezidis HD, Adamopoulos AD, Kassomenos PA. Comparison between experimental data and modeling estimates of aerosol optical depth over Athens, Greece. Journal of Atmospheric and Solar-Terrestrial Physics 2006;68:1167–78.
[53] Bird Hulstrom RL. Direct insolation models. Transactions of ASME – Journal of Solar Energy Engineering 1981;103:182–92.
[54] Bird RE, Hulstrom RL. A simplified clear sky model for direct and diffuse insolation on horizontal surfaces. SERI/TR; 1981. p. 642–761.
[55] Melice JL, Boughanmi A, Eaton E, Wendler G. Turbidity measurements of Saharan aerosols and their effect on atmospheric heating and planetary reflectivity. Archives for Meteorology, Geophysics, and Bioclimatology Series B 1984;35:203–20.
[56] Nicholls RW. Wavelength-dependent spectral extinction of atmospheric aerosols. Applied Optics 1984;23:1142–3.
[57] Bird RE. A simple solar–spectral model for direct-normal and diffuse horizontal irradiance. Solar Energy 1984;32:461–71.
[58] Fox Jd. Calculated Ångström's turbidity coefficients for Fairbanks, Alaska. Journal of Climate 1994;7:1506–12.
[59] Kaskaoutis DG, Kambezidis HD, Adamopoulos AD, Kassomenos PA. On the characterization of aerosols using the Ångström exponent in the Athens area. Journal of Atmospheric and Solar-Terrestrial Physics 2006;68:2147–63.
[60] O'Neill N, Royer A. Extraction of bimodal aerosol-size distribution radii from spectral and angular slope (Angstrom) coefficients. Applied Optics 1993;32:1642–5.
[61] Kaufman YJ. Aerosol optical thickness and atmospheric path radiance. Journal of Geophysical Research 1993;98(D2):2677–92.
[62] Gueymard C. Development and performance assessment of a clear sky spectral radiation model. In: Proc. Solar'93-22 d ASES Conf. 1993. p. 433–8.
[63] Gueymard C. Updated transmittance functions for use in fast spectral direct beam irradiance models. In: Proc. 23d ASES Conf. 1994. p. 1–6.
[64] Kaufman YJ, Gitelson A, Karnieli A, Ganor E, Fraser RS, Nakajima T, et al., Holben BN. Size distribution and phase function of aerosol particles retrieved from sky brightness measurements. Journal of Geophysical Research 1994;99(10):10341–56.
[65] Kasten F. The Linke turbidity factor based on improved values of the integral Rayleigh optical thickness. Solar Energy 1996;56(3):239–44.
[66] Bokoye AI, Casinière AL, Cabot T. Ångström turbidity parameters and aerosol optical thickness: a study over 500 solar beam spectra. Journal of Geophysical Research 1997;102(21):905–14.
[67] Eck TF, Holben BN, Reid JS, Dubovik O, Smirnov A, O'Neill NT, et al. Wavelength dependence of optical depth of biomass burning, urban and desert dust aerosols. Journal of Geophysical Research 1999;104(D24):31333–50.
[68] Malik AQ. A modified method of estimating Ångström's turbidity coefficient for solar radiation models. Renewable Energy 2000;21:537–52.
[69] Adeyewa ZD, Balogun EE. Wavelength dependence of aerosol optical depth and the fit of the Ångström law. Theoretical and Applied Climatology 2003;74:105–22.
[70] Adeyefa ZD, Holmgren B, Adedokun JA. Spectral solar irradiance in Northern Scandinavia before and after Pinatubo. Atmosfera 2000;13(3):13346.
[71] O'Neill NT, Dubovik O, Eck TF. A modified Ångström coefficient for the characterization of sub-micron aerosols. Applied Optics 2001;40(14):2368–75.
[72] Schuster GL, Dubovik O, Holben BN. Ångström exponent and bimodal aerosol size distributions. Journal of Geophysical Research 2006;111(D07207), doi:10.1029/2005JD006328.
[73] Gobbi GP, Kaufman YJ, Koren I, Eck TF. Classification of aerosol properties derived from AERONET direct sun data. Atmospheric Chemistry and Physics 2007;7(2):453–8.
[74] Cucumo M, Marinelli V, Oliveti G. Experimental data of the Linke turbidity factor and estimates of the Ångström turbidity coefficient for two Italian localities. Renewable Energy 1999;17:397–410.
[75] Wagner F, Silva AM. Some considerations about Ångström exponent distributions. Atmospheric Chemistry and Physics 2008;8:481–9.
[76] Kaskaoutis DG, Kambezidis HD. Checking the validity of the Ångström's formula with spectral data of direct-beam irradiance obtained in Athens, Greece. Atmospheric Research 2006;79:67–87.
[77] Polo J, Zarzalejo LF, Salvador P, Ramıŕez L. Ångström turbidity and ozone column estimations from spectral solar irradiance in a semi-desertic environment in Spain. Solar Energy 2009;83:257–63.
[78] Pedrós R, Utrillas MP, Martínez-Lozano JA, Tena F. Values of broad band turbidity coefficients in a Mediterranean coastal site. Solar Energy 1999;66(1):11–20.
[79] Cañada J, Pinazo JM, Bosca JV. Determination of turbidity coefficient at Valencia. Renewable Energy 1993;3(6–7):621–6.
[80] Cachorro VE, Vergaz R, de Frutos AM. A quantitative comparison of α Ångström turbidity parameter retrieved in different spectral ranges based on spectroradiometer solar radiation measurements. Atmospheric Environment 2001;35:5117–24.
[81] Ganesh KE, Umesh TK, Narasimhamurthy B. Atmospheric turbidity over a continental station Mysore, India. Indian Journal of Radio & Space Physics 2011;40:85–94.
[82] Balakrishnaiah G, Kumar KR, Reddy BSK, Gopal KR, Reddy RR, Reddy LSS, Ahammed YN, Narasimhulu K, Moorthy KK, Babu SS. Analysis of optical properties of atmospheric aerosols inferred from spectral. Atmospheric Environment 2011:1275–85.
[83] Pinker RT, Pandithurai G, Holben BN, Keefer TO, Goodrich D. Aerosol radiative properties in the semiarid Western United States. Atmospheric Research 2004;71:243–52.
[84] Sapkota B, Dhaubhadel R. Atmospheric turbidity over Kathmandu valley. Atmospheric Environment 2002;36:1249–57.
[85] Salazar GA. Estimation of mo nthly values of atmosphe ric turbidity using measured values of global irradiation and estimated values from CSR and Yang Hybrid models, Study case: Argentina. Atmospheric Environment 2011;45:2465–72.
[86] Tadros MTY, El-Metwally M, Hamed AB. Determination of Ångström coefficients from spectral aerosol optical depth at two sites in Egypt. Renewable Energy 2002;27:621–45.
[87] Maduekwe AAL, Chendo AC. Atmospheric turbidity and the diffuse irradiance in Lagos Nigeria. Solar Energy 1997;61(4):241–9.
[88] Trabelsi A, Masmoudi M. An investigation of atmospheric turbidity over Kerkennah Island in Tunisia. Atmospheric Research 2011;101:22–30.
[89] Dogniaux R. Representation analytique des composantes du rayonnement solaire. Institut Royal de Météorologie de Belgique, Série A; 1974. No. 83.

[90] Molineaux B, Ineichen P, Delaunay JJ. Direct luminous efficacy and atmospheric turbidity improving model performance. Solar Energy 1995;55(2):125–37.
[91] Kasten F. A simple parameterization of two pyrheliometric formulae for determining the Linke turbidity factor. Meteorology Rdsch 1980;33:124–7.
[92] CSAGI (Special Committee for the International Geophysical Year). Annals of the International Geophysical Year, Vol. V, Part VI-Radiation instruments and measurements, pp. 404 and 463. Springer, London (1958).
[93] Ineichen P, Perez R. A new air mass independent formulation for the Linke turbidity coefficient. Solar Energy 2002;73(3):151–7.
[94] Louche A, Peri G, Iqbal M. An analysis of Linke turbidity factor. Solar Energy 1986;37:393–6.
[95] Grenier JC, De La Casiniere A, Cabot T. Atmospheric turbidity analyzed by means of standardized Linke's turbidity factor. Journal of Applied Meteorology 1995;34:1449–58.
[96] Aydinli S. The Availability of Solar Radiation and Daylight, Draft Technical Report to CIE Technical Committee 4.2, October 1981.
[97] Abdelrahman MA, Said SAM, Shuaib AN. Comparison between atmospheric turbidity coefficients of desert and temperate climates. Solar Energy 1988;40:219–25.
[98] Dogniaux R, Page JK. European Solar Radiation Atlas (ESRA), includ. CD-ROM. In: Greif J, Scharmer K, Scientific advisors. Published for the Commission of the European Communities. Wald L, Albuisson M, Czeplak, G, Bourges B, Aguiar R, Lund H, Jouko A, Terzenbach U, Beyer HG, Borisenko EP, editors. Ecole des Mines de Paris, France: Presses de l'Ecole; 2000.
[99] Cucumo M, Kaliakatsos D, Marinelli V. A calculation method for the estimation of the Linke turbidity factor. Renewable Energy 2000;19:249–58.
[100] ESRA, 1984. European Solar Radiation Atlas. Second Improved and Extended Edition, vols. I and II (1984). In: Palz, W. (Ed.), Commission of the European Communities, DG Science, Research and Development, Report No. EUR 9344, Brussels.
[101] Walkenhorst O, Luther J, Reinhart C, Timmer J. Dynamic annual daylight simulations based on one-hour and one-minute means of irradiance data. Solar Energy 2002;72(5):385–95.
[102] Mavromatakis F, Franghiadakis Y. Direct and indirect determination of the Linke turbidity coefficient. Solar Energy 2007;81:896–903.
[103] Chaiwiwatworakul P, Chirarattananon S. An investigation of atmospheric turbidity of Thai sky. Energy and Buildings 2004;36:650–9.
[104] Rapti AS. Atmospheric transparency, atmospheric turbidity and climatic parameters. Solar Energy 2000;69(2):99–111.
[105] Coste L, Eftimie E. Linke turbidity modelling for Brasov urban area. In: International Conference on Renewable Energies and Power Quality (ICREPQ'10). 2010.
[106] Kryza M, Szymanowski M, Migała K, Pietras M. Spatial information on total solar radiation: Application and evaluation of the r.sun model for the Wedel Jarlsberg Land, Svalbard. Pol Polar Res 2010;31(1):17–32.